To main QEN pages

To new maps homepage

Article in press in *Progress in Physical Geography*

This represents an earlier version of our text. Some changes have been made since we stopped modifying this web version: e.g. we have added a discussion of the role of volcanic aerosols in sudden climate changes...evidence suggests the rapid cooling at the end of the Eemian interglacial was due to a big explosive volcanic event. Other 'volcanic' cooling events occured during the Holocene.

Sudden climate transitions during the Quaternary

by Jonathan Adams (1.), Mark Maslin (2.) & Ellen Thomas (3.)

(1.) MS 6335, Environmental Sciences Division, Oak Ridge National Laboratory, Oak Ridge, TN 37831, USA

(2.) Environmental Change Research Centre, Department of Geography, University College London, 26 Bedford Way, London, WC1H 0AP, UK

(3.) Center for the Study of Global Change, Department of Geology and Geophysics, Yale University, PO Box 208109, New Haven, Connecticut 06520-8109, USA *and* **Department of Earth and Environmental Sciences, Wesleyan University, 265 Church Street, Middletown CT 06459-0139, USA.**

Abstract

The time span of the past few million years has been punctuated by many rapid climate transitions, most of them on time scales of centuries to decades or even less. The most detailed information is available for the Younger Dryas-to-Holocene stepwise change around 11,500 years ago, which seems to have occurred over a few decades. The speed of this change is probably representative of similar but less well-studied climate transitions during the last few hundred thousand years. These include sudden cold events (Heinrich events/stadials), warm events (Interstadials) and the beginning and ending of long warm phases, such as the Eemian interglacial. Detailed analysis of terrestrial and marine records of climate change will, however, be necessary before we can say confidently on what timescale these events occurred; they almost certainly did not take longer than a few centuries.

Various mechanisms, involving changes in ocean circulation, changes in atmospheric concentrations of greenhouse gases or haze particles, and changes in snow and ice cover, have been invoked to explain these sudden regional and global transitions. We do not know whether such changes could occur in the near future as a result of human effects on climate. Phenomena such as the Younger Dryas and Heinrich events might only occur in a 'glacial' world with much larger ice sheets and more extensive sea ice cover. However, a major sudden cold event did probably occur under global climate conditions similar to those of the present, during the Eemian interglacial, around 122,000 years ago. Less intensive, but significant rapid

climate changes also occurred during the present (Holocene) interglacial, with cold and dry phases occurring on a 1500-year cycle, and with climate transitions on a decade-to-century timescale. In the past few centuries, smaller transitions (such as the ending of the Little Ice Age at about 1650 AD) probably occurred over only a few decades at most. All the evidence indicates that most long-term climate change occurs in sudden jumps rather than incremental changes.

Introduction

Until a few decades ago it was generally thought that all large-scale global and regional climate changes occurred gradually over a timescale of many centuries or millennia, scarcely perceptible during a human lifetime. The tendency of climate to change relatively suddenly has been one of the most suprising outcomes of the study of earth history, specifically the last 150,000 years (e.g., Taylor et al., 1993). Some and possibly most large climate changes (involving, for example, a regional change in mean annual temperature of several degrees celsius) occurred at most on a timescale of a few centuries, sometimes decades, and perhaps even just a few years. The decadal-timescale transitions would presumably have been quite noticeable to humans living at such times, and may have created difficulties or opportunities (e.g., the possibility of crossing exposed land bridges, before sea level could rise). Hodell et al. (1995) and Curtis et al. (1996), for instance, document the effects of climate change on the collapse of the Classic period of Mayan civilization and Thompson (1989) describes the influence of alternating wet and dry periods on the rise and fall of coastal and highland cultures of Ecuador and Peru. The beginning of crop agriculture in the Middle East corresponds very closely in time with a sudden warming event which marks the beginning of the Holocene (Wright 1993). Even the burial in ice of the prehistoric mummified corpse of the famous 'Iceman' (e.g., Bahn and Everett, 1993) at the upper edge of an alpine glacier coincided with the initiation of a cold period ('Neoglaciation') after the Holocene climate optimum (Baroni and Orombelli, 1996). On longer timescales, evolution of modern humans has been linked to climatic changes in Africa (e.g., de Menocal, 1995). But the full implications of these sudden changes for biogeography and for the evolution of human cultures and biology have barely begun to be considered; there has simply not been time for the message to be absorbed by biogeographers, archaeologists and palaeoanthropologists, and this review is intended to help the process along.

Sudden stepwise instability is also a disturbing scenario to be borne in mind when considering the effects that humans might have on the climate system through adding greenhouse gases. Judging by what we see from the past, conditions might gradually be building up to a 'break point' at which a dramatic change in the climate system will occur over just a decade or two, as a result of a seemingly innocuous trigger. It is the evidence for dramatic past changes on the timescale of centuries to decades which will be the subject of this review.

I. Broad climate variability; the background of oscillations on the timescale of tens of thousands of years

Climatic variability on the timescale of tens of thousands of years has turned out to be a predominant pattern in earth history. The last two and a half million years have been marked by many global climate oscillations, between warmer and cooler conditions. This trend of oscillations appears to be merely the continuation of a pattern of variability extending back well into the Tertiary period and possibly beyond (e.g., Kennett 1995). During the last few million years, the length and the amplitude of these climate cycles has increased (e.g., Crowley & North, 1991; Hodell and Venz, 1992).

Large global interglacial-glacial-interglacial climate oscillations have been recurring at approximately a 100,000 year periodicity for the last 900,000 years (e.g. Berger et al. 1993; Mudelse and Schulz, 1997), though each individual cycle has had its own idiosyncrasies in terms of the timing and magnitude of changes (e.g., Lyle et al. 1992). As is usually the case with the study of the past, data become more scarce with increasing age (Winograd et al. 1997); even so, many detailed records are now becoming available (e.g. Tzedakis et al 1997). Extended records of atmospheric gas concentrations and polar temperatures may also be expected from the continued deeper drilling of the Antarctic Vostok ice core (Jouzel et al. 1996, Petit et al. 1997).

The most recent large climate oscillation spanning the last 130,000 years (130 ka) has been the subject of the most intensive study, because it offers a relatively detailed climate record from the land, from the oceans and from the ice cores. In the last few years, a considerable amount of new data on the warm period known as the Eemian (the last major interglacial) has become available (e.g. Pewe et al. 1997). This interval has seen the global climate system switch (Fig.1) from warm interglacial (similar to present-day) to cold glacial conditions, and back.

In stratigraphic terms, the last 100 ka oscillation began with the Eemian interglacial (Isotope Stage 5e, possibly including part of the series 5d-a; known as the Sangamon Stage on land in the USA), followed by the colder last glacial period (Stages 4-2), and the present Holocene Interglacial beginning around 11,000 years ago (Isotope Stage 1). There is considerable evidence that similar events and processes were at work in previous glacial-interglacial cycles over the past 900,000 years. The general background of continuous (but less dramatic) variability in the earth's climate system is thought to extend well back into the Tertiary, and beyond (e.g., Crowley and North 1991).

For the most recent Eemian-to-Holocene phase there is still significant ambiguity in terms of the errors in geological dating techniques, causing difficulties in correlation between different regions, so that it is difficult to know whether climate changes were truely synchronous globally. However, in many other respects the record that has been assembled for this period is remarkably detailed. There are apparently few gaps in the record among the many long ocean cores, though land and lake records are much patchier. Indicator species on land (e.g., trees) may respond slowly, but oceanic unicellular plankton (such as diatoms, foraminifera and nannoplankton) has such short generation times that it can potentially give a precise picture of events, at least where sedimentation rates are sufficient. For instance, exciting new data can be expected from detailed analysis of the cores taken during Leg 172 of the international Ocean Drilling Program of 1997, which concentrated on high sedimentation rate sediment drifts in the North Atlantic.

The general picture summarized here roughly reflects the present consensus gained from studies of ice cores, deep ocean cores, and terrestrial and lake sediments around the world. This consensus is itself subject to sudden jumps when new data are presented, or as more thorough reanalyses of previous data come forth; for this reason, our review is liable to be significantly out of date within only a few weeks or months of being written!

II. The record of decade-to-century timescale changes during the last 130,000 years

II.1. The Eemian, or last interglacial.

The last interglacial (also called the Eemian) has often been seen as a close counterpart of the present interglacial: sea surface temperatures were similar, and sea level was possibly somewhat higher (e.g., Imbrie and Imbrie, 1992). Going by the principle that there is a general similarity between these two warm periods, the Eemian has been used to predict the duration of the present interglacial, and also to study the possibility of sudden climate variability occurring within the next few centuries or millennia. The dating of the warming event that begins the Eemian is still the subject of controversy, and this leads to greater uncertainty about what to make of it as an analogue for future climate change.

II.1.a. Controversy over the timing of the last interglacial. The Eemian or Marine oxygen Isotope Stage (MIS) 5e interglacial (Fig.1) began sometime between 130-140 ka ago (Imbrie et al. 1984; Martinson et al., 1987; Sarnthein & Tiedemann, 1990; Szabo et al., 1994; Stirling et al., 1995) with a warming phase (of uncertain duration) taking the earth out of an extreme glacial phase, into conditions generally warmer than those of today (Frenzel et al. 1992). Warming into the Eemian may have occurred in two major steps, similar to the last deglaciation (Seidenkrantz et al., 1993; 1996).

Though it was named after a warm-climate phase seen in the terrestrial pollen record of the Netherlands (e.g. Zagwijn, 1963, 1975), the first generally accepted dates of the start of the Eemian came from the oceans (e.g. Shackleton 1969), based on interpretation of benthic foraminiferal oxygen isotope data. Such oxygen isotope curves were later expanded and many curves were stacked; the timing of maximum ice volume as estamited from oxygen isotope values was then 'tuned' to astonomical variations in solar input (see below). This time scale is usually called the called the SPECMAP curve (Imbrie et al. 1984; Martinson

et al., 1987). The warm MIS 5e, the warmest interval in the last 150,000 years, was correlated with the warmest interval in pollen records, the Eemian. The land records were then dated by 14C for the younger parts of the records, by correlation with the marine records for the older parts (e.g., Woilard, 1979; Woillard and Mook, 1982).

The age of the warming event (or events) at the beginning of stage 5e, however, is still under discussion. Work on deep-sea sediments (Imbrie et al., 1993; Sarnthein and Tiedemann, 1990; Maslin et al., 1998) and corals (Szabo et al., 1994; Stirling et al., 1995; Slowey et al., 1996) suggests that rapid warming could have started as early as 132 ka, while work on the Antarctic Vostok ice core suggests a possible initiation at 134 kyr (Jouzel et al., 1993, Nature 364, 407-411). Uranium-Thorium dated records from a continental karst sediment in the southwestern USA (Devil's Hole; Winograd et al., 1988; 1992; 1997), however, suggest a much earlier start of warming at about 140 kyr. Studies of an Alaskan site (the Eve Interglaciation Forest Bed) suggest, however, that the Devil's Hole record may be in error and that the warming definitely postdated 140 kyr, because the tephra underlying this bed was dated at 140 kyr. There has been much discussion about the reliablity of dating of the marine isotope stages as compared to the Devil's Hole record (e.g., Imbrie et al., 1993), but new numerical dating methods have apparently confirmed that both records are reliable (Protactinium-231 dating of carbonates; Edwards et al., 1997). We must thus consider the possibility that warm conditions did not last for the same amount of time throughout the world; for instance comparison of various land records also suggests that warming may have occurred at different times in the Alps and in northern France (de Beaulieu and Reille, 1989).

Winograd et al. (1997) date the ending of the warm period at about the same time as the SPECMAP time scale, so that they require the duration of the Eemian warm period to have been about twice as long as in the SPECMAP scheme of events (i.e. lasting 22 kyr rather than about 10 kyr). We need to consider the possibility that warm intervals as seen in pollen records have a longer duration than periods of high sea level and low ice volume in the marine record: Kukla et al., 1997, for instance, suggest that the Eemian (as seen in the pollen records) started at about 130 kyr, but ended much later than the end of MIS 5e, and that the duration of land interglacials thus is indeed longer than the period of low ice volume. Such a difference in duration is also apparent in the comparison of terrestrial and land records by Tzedakis et al., 1997 (their fig. 2).

If one accepts the contradictory picture obtained from comparing the Devil's Hole record and some of the terrestrial pollen records with other parts of the world, it seems that for thousands of years warm 'interglacial' type conditions in the mid-latitudes on land could have been occurring at the same time as much colder ocean temperatures and expanded Arctic ice sheets. The Eemian, it appears, could have been a strange beast quite unlike our present interglacial phase (which began rapidly and fairly simultaneously all around the world). This confusion over the nature and duration of the Eemian adds to the difficulty in making simple, general comparisons with our present interglacial, and in interpreting the significance of some of the events seen in the marine and ice core records.

II.1.b. Evidence of climate instability during the Eemian

Whatever the true timescale of the Eemian, what has generated most widespread interest (extending well beyond the usual geological community and into the popular media) is that there are indications of large-scale climate instability in the middle of the Eemian (e.g., Maslin and Tzedakis, 1996). This finding is alarming, as we are presently living in an interglacial period (called the Holocene) which closely resembles the Eemian in many respects (though it is not yet clear how closely; Imbrie and Imbrie, 1992; fig. 48). If sudden, dramatic climate changes could occur within the Eemian, then they could perhaps occur in the future during our present interglacial, especially if we perturb the system by adding greenhouse gases.

Initial evidence from the GRIP ice core evidence (Dansgaard et al., 1993; Taylor et al. 1993) indicated that the Eemian was punctuated by many short-lived cold events, as shown by variations in electrical conductivity (a proxy for windblown dust, with more dust indicating colder, more arid conditions) and stable oxygen isotopes (a proxy for air temperature) of the ice were used by these workers infer the climatic conditions during the Eemian. The cold events seemed to last a few thousand years, and the magnitude of cooling was similar to the difference between glacial and interglacial conditions; a very dramatic contrast in climate. Furthermore, the shifts between these warm and cold periods seemed to be extremely rapid,

possibly occurring over a few decades or less.

A second ice core (GISP2) from the Greenland ice cap (Boulton, 1993) provided an almost identical climate record for the last 110 ka, shortly after the end of the Eemian. GISP2 also shows apparent sudden climate jumps throughout the Eemian, but the two records diverge (Grootes et al. 1993). Significantly, in GISP2 steeply inclined ice layers occur in this lower portion of the core, indicating that the ice has been disturbed, and that we cannot distinguish simple tilting from folding or slippage that would juxtapose ice of very different ages (Boulton, 1993). For this reason. the deeper GISP2 record has thus been interpreted as containing interglacial and glacial ice of indeterminate age, mixed in together by ice tectonics (Grootes et al., 1993). It has been suggested that the deeper parts of the GRIP ice core record (referred to above), including the crucial Eeemian sequence, may also have been affected by ice tectonics (e.g. Grootes et al., 1993; Taylor et al., 1993; Boulton, 1993). Johnsen et al. (1995) reported layers tilted up to 20 degrees within the Marine Isotope Stage 5c (110 ka) section of the GRIP ice cores, precisely where the correlation between GRIP and GISP2 breaks down. The GISP2-GRIP Joint Workshop in Wolfeboro, New Hampshire, USA (September 1995), concluded from data on ice trapped atmospheric CH4 and d18O from both Antartctica and Greenland, and from more detailed work on the structural properties of the cores, that both GRIP and GISP2 ice cores had suffered stratigraphic disturbances in ice older than 110 ka. There are presently plans to obtain high resolution last interglacial records from both Greenland and the high accumulation rate West Antarctic ice sheet, which will add an interhemispheric dimension to the current Eemian debate.

Support for the occurrence of cold Eemian events was obtained from lake records from continental Europe (de Beaulieu & Reille 1989, Guiot et al. 1993), the Massif Central in France (Thouveny et al., 1994) and Bispingen, Germany (Field et al., 1994). However, these records only suggest that the Eemian climate was more variable than that of the Holocene, not that it had the extreme frenetic variability of glaical times that the ice cores suggest. Oceanic records from the North East Atlantic (McManus et al., 1994) and the Bahamas Outer Ridge (Keigwin et al., 1994) indicate very little or no climatic variability during the Eemian. In contrast to this, records from both the Nordic Seas and west of Ireland show a cooling and freshening of the North Atlantic in the middle of the Eemain somewhere between 122 and 125 ka (Cortijo et al., 1994; Fronval and Jansen, 1996). These Nordic records also show highly variable surface water conditions throughout the Eemian period. Records from slightly further south on the north-west European shelf sediments, suggest a similar picture of cold intervals during the Eemian (Seidenkrantz et al., 1995).

Evidence for a single sudden cool event during the Eemian is also present in pollen records from a lake in central Europe (Field et al., 1994), from loess sedimentology in central China (Zhisheng & Porter 1997), and from ocean sediment records from ODP Site 658 in the eastern sub-tropical Atlantic (Maslin et al., 1996; 1998; Maslin and Tzedakis, 1996). However, although widespread, it may not have been a global event, as it apparently does not show up in cores from some parts of the world (e.g., Oppo et al.1997).

Overall, the combined sources of evidence suggest that at least one cold and dry event caused a change in circulation patterns in the North Atlantic, a several-degree decline in Atlantic surface temperatures, and on land opened up the west European forests to give a mixture of steppe and trees near the middle of the Eemian, about 122 ka. This intra-Eeemian cold phase was less dramatic than had been suggested by the variability in the ice core records, but it was still a major climatic change. Evidence from a high resolution marine core record at Site ODP 658 (Maslin and Tzedakis, 1996) suggests that this event, which might possibly have come on in a few decades or less, lasted no more than 400 years. Afterwards climate recovered, but conditions did not return to the full warmth of the early Eemian 'optimum'.

Further support for the existence of the intra-Eemian cooling event comes indirectly from coral reef records. New data from Stirling et al. (1995) from Western Australia and their review of other high precision U-series coral dates indicate that during the last interglacial period (which they date to between 130 to 117 ka) the main global episode of coral reef building was confined to just a few thousand years between 127 to 122 ka. This period of reef building thus seems to have ended at the same time as the beginning of the Intra-Eemian cold event at about 122 ka. This intra-Eemian event seems to have have a corresponding, though admittedly weaker, equivalent in the present interglacial (the Holocene). Around 4,800-4,500 ya, there was a 300-year long colder phase that resembled the Intra-Eemian cooling event (Fig. 3) in that they both occur after the interglacial peak and signal the beginning of a trend of climatic deterioration (e.g., Baroni and

Orombelli, 1996; Bond et al., 1997)

Other large and relatively sudden cool and arid phases (occurring against a background of similar-to-present conditions) seem to have affected some of the previous interglacials before about 200,000 years ago (Winograd et al. 1997). Again, the speed with which these climate transitions occurred is unclear through lack of detailed time resolution in these older records, but the possibility that these changes occurred over only a few decades must be considered a possibility.

II.2. Sudden transitions after 115,000 years ago.

According to the marine records, the Eemian interglacial ended with a rapid cooling event about 110,000 years ago (e.g., Imbrie et al., 1984; Martinson et al., 1987), which also shows up in ice cores and pollen records from across Eurasia. From a relatively high resolution core in the North Atlantic. Adkins et al. (1997) suggested that the final cooling event took less than 400 years, and it might have been much more rapid.

Following the end of the Eemian, a large number of other sudden changes and short-term warm and cold alternations have been recognized; apparently many or all of these occurred on a global or at least a regional scale (Fig.3; Ice core record). The most extreme of these fluctuations are the warm interstadials and the cold Heinrich events. These are most prominent in the ice-core record of Greenland, deep-sea cores from the North Atlantic, and in the pollen records of Europe and North America, suggesting that they were most intense in the North Atlantic region (e.g., Bond et al., 1992; 1993).

II.2.a. Interstadials.

Sudden and short-lived warm events occurred many times during the generally colder conditions that prevailed between 110,000 and 10,000 years ago (isotope Stages 2-5.4). First picked up as brief influxes of warm climate plants and insects into the glacial tundra zone of northern Europe, they are known as 'interstadials' to distinguish them from the cold phases or 'stadials' (Lowe & Walker 1984). The interstadials show up strongly in the Greenland ice core records. Between 115,000 and 14,000 years ago, 24 of these warm events have been recognized in the Greenland ice cores (where they are called 'Dansgaard-Oeschger events'; e.g., Bond et al., 1993; Bond and Lotti, 1995). Many lesser warming events have also been seen in the ice core records (e.g., Dansgaard et al., 1993; Taylor et al., 1993; 1997; Mayewski et al., 1997) but have not yet been recognized elsewhere. Short-lived and/or moist warm phases, coeval with interstadials, appear in the eastern Pacific (Behl and Kennett 1996), western Siberia, the Arabian Sea (Schulz et al., 1998) and possibly also in central China (Behl & Kennett 1996). The duration of each interstadial in the can be counted in ice cores from the annual 'snowlayers that have accumulated for the younger parts of the records, or (rather less precisely) from the thickness of sediment accumulated in an ocean bed core. In marginal ocean basins with low oxygenation, annual layers may be preserved in the sediment (e.g., Behl and Kennett, 1996; Ocean Drilling Program Leg 169S, Saanich Inlet, British Coumbia, Canada). Ice core and ocean data suggest that interstadials both began and ended suddenly, though in general the 'jump' in climate at the start of an interstadial was followed by a more gradual decline involving a stepwise series of smaller cooling events and often a fairly large terminal cooling event which returned conditions to the colder 'glacial' state (e.g., Rasmussen et al., 1997). From the ice core evidence from Greenland, warming into each interstadial occurred over a few decades or less, and the overall duration of some of these warm phases may have been just a few decades, while others vary in length from a few centuries to nearly two thousand years (e.g, Mayewski et al., 1997).

II.2.b.Heinrich events.

Opposite in sign to the Interstadials are extreme and short-lived cold events, known as 'Heinrich events' (e.g., Heinrich, 1988; Bond et al., 1992; Grousset et al., 1993; Andrews et al., 1994), which were first recognized as periods with very ice-rafting in the North Atlantic (Ruddiman, 1977). These events occur against the general background of the glacial climate, and represent the brief expression of the most extreme glacial conditions. The several massive ice-rafting events (e.g., Heinrich 1988; Broecker et al., 1992; Bond et al., 1992; 1993; Maslin et al., 1995) (Table 1) also show up in the Greenland ice cores as a further 3-6 deg.C drop in temperature from the already cold glacial climate (Bond et al., 1993; Mayewski et

al., 1994), and many of these events also been picked up as particularly cold and arid intervals in European and North American pollen records (Grimm et al., 1993). There is still considerable discussion as to the exact region of provenance of the icebergs participating in the Heinrich events (e.g., Gwiazda et al., 1996a, b; Revel et al., 1996; Lehman, 1996; Bond et al., 1997). At North Atlantic mid-latitudes, oceanic surface productivity dropped precipitously during Heinrich events (Thomas et al., 1995). It is thought that at least some of these events also affected the climate further afield from Greenland - giving cold, arid conditions as far away as central China and Antarctica (Thompson et al., 1989). Preliminary data (C. Whitlock..web page) from marine sediments off California and Oregon, pollen records from Pacific Northwest lakes, cores from the Arabian Sea (Schulz et al. 1998) and glacial records from western North America all reveal climate events that appear to be related to the Heinrich events of the North Atlantic. Thus it appears that these may have been global events, although the climate shift may have been smaller outside the North Atlantic region.

The last Heinrich event (known as H1) *sensu stricto* occurred just after the Last Glacial Maximum and seems to mark the extreme cold and aridity that occurs in many parts of the world around 17,000-15,000 years ago. The Younger Dryas cold phase (see below) may also be regarded as a 'Heinrich' climate event (it is sometimes now referred to as H0), and as it has been studied in considerable detail it may give clues to a general pattern true to all Heinrich climate events. The detailed timescale on which most of these ice rafting and climate change events began and ended is uncertain, though it cannot have been very much longer than several decades because the events themselves lasted no more than a few centuries (e.g., Francois and Bacon, 1993; Mayewski et al., 1994; Dowdeswell et al., 1995). By analogy with the well-studied Younger Dryas event (below), Heinrich events may be regarded as possibly beginning and ending with sudden climate 'jumps' taking just a few decades. However, this idea remains tentative without further work.

Recently, detailed studies of the sequence of events in ocean sediments and ice cores shows that Heinrich events actually occurred more frequently around the rim of the North Atlantic - about every 7,000-10,000 years on average, in the time interval between about 70,000 and 10,500 years ago. In the later stage of this time span, after about 38,000 y.a., they were even more frequent at about every 1,000-3,000 years, apparently following a mean cycle length of about 1500 years (Bond et al. 1997). Similar but lower amplitude 1500-year oscillations have been found to occur during the present Holocene interglacial (Campbell et al., 1998; see below) as well as during earlier glacial and interglacial stages (Oppo et al. 1998).

II.2.c. The Younger Dryas.

The Younger Dryas cold event at about 12,900-11,500 years ago seems to have had the general features of a Heinrich Event, and may in fact be regarded as the most recent of these (Severinghaus et al. 1998). The sudden onset and ending of the Younger Dryas has been studied in particular detail in the ice core and sediment records on land and in the sea (e.g., Bjoerck et al., 1996), and it might be respresentative of other Heinrich events. A detailed study of two Greenland ice cores (GRIP and GISP2; Taylor et al. 1997), suggests that the main Younger Dryas-to-Holocene warming took several decades in the Arctic, but was marked by a series of warming steps, each taking less than 5 years. About half of the warming was concentrated into a single period of less than 15 years. A rapid global rise in atmospheric methane concentration which occurred at the same time suggests that the warming and moistening of climate (causing more methane output from swamps and other biotic sources) was a globally synchronized change (Meeker et al.1997; Fuhrer and Legrand, 1997). According to data from the Greenland ice-cores, conditions remained slightly cooler than present for a while after the main warming period; 'normal' Holocene warmth was not reached for a further 1500 years (up until around 10,000 calendar years ago). It is not yet clear if the general pattern of the transition between the Younger Dryas and Holocene is representative of other rapid warming and cooling events in the past 110,000 years, but similar events seem to have occurred at the beginning of the Eemian (Seidenkrantz et al. 1995).

Other possibly sudden climate transitions since the start of the Holocene. Following the sudden start of the Holocene about 11,500 years ago, there have been a number of rapid, widespread climate changes recorded from the palaeoclimatic record around the world. The Greenland ice core data again show a clear record of these events (O'Brien et al., 1996; Mayewski et al., 1997).

At least in the North Atlantic region, these changes seem to have been paced according to approximately the same 1500-year rhythm as that found for the last glacial and earlier glacial periods, according to Atlantic sediment records (Bond et al. 1997; Campbell et al., 1998). Generally, at the coldest point of each 1500-year cycle surface temperatures of the North Atlantic were about 2 deg.C cooler than at the warmest part, representing a fairly substantial change in climate. Regional or global fluctuations of this order would be major events if they were to suddenly affect the present-day world with its high population and finely balanced food production. It is uncertain whether these climate cycles indeed extended around the world or were generally confined to the region around the North Atlantic, but the 8,200 ka event (see below) (which fits in as one of the more extreme cold events of this 1500-year pattern) does seem to have been widespread. Other veents might be recognizable as variations in monsoonal intensity in the Indian Ocean, but there is still uncertainty in exact time correlations (Sirocko et al., 1993).

The event at 8200 ka is the most striking sudden cooling event during the Holocene, giving widespread cool, dry conditions lasting perhaps 200 years before a rapid return to climates warmer and generally moister than the present. This event is clearly detectable in the Greenland ice cores, where the cooling seems to have been about half-way as severe as the Younger Dryas-to-Holocene difference (Alley et al., 1997; Mayewski et al., 1997). No detailed assessment of the speed of change involved seems to have been made within the literature (though it should be possible to make such assessments from the ice core record), but the short duration of these events at least suggests changes that took only a few decades or less to occur. Coeval records from North Africa across Southern Asia, show markedly more arid conditions involving a failiure of the summer monsoon rains. Cold and/or aridity also seems to have hit northernmost South America, eastern North America and parts of NW Europe (Alley et al., 1997). Thinking of our densely populated present-day world, we can only hope that no similar event occurs in the near future. Smaller, but also sudden and widespread, changes to drier or moister conditions have also been noted for many parts of the world for the second half of the Holocene, since about 5,000 years ago (e.g., Dorale et al., 1992). One fairly strong arid event occurred about 4,000 years ago across northern Africa and southern Asia. It remains to be seen whether these later events will eventually fit into a consistant global 1500-year pattern of cold/arid events, the last of which may have been the Little Ice Age which ended about 350 years ago.

The effects which mid-Holocene climate fluctuations might have had on regional ecology are still being worked out. In Holocene pollen records, this is the period when there was an elm (*Ulmus*) decline in Europe (about 5,000 14C yrs or 5700 calendar yrs) and the hemlock (*Tsuga*) decline in North America (about 4,700 4C yrs or 5300 calendar yrs). Both of these have been attributed to specific pathogen attacks (Rackham, 1980; Peglar 1993), but with the evidence from the Eemian it may now be worth considering these declines in terms of climate deterioration, or at least its effect on the spread of epidemics (A. Parker pers. comm.)

II.2.d. Other sudden climate jumps from the more distant past

Climate instability on the timescale of tens of thousands of years has been found in deep ocean and lake cores going back more than 40 Myr (e.g., Crowley and North, 1991). However, because these deeper records have a relatively poor time resolution, it is uncertain whether sudden decade-century timescale jumps in climate were common before the Quaternary Period (the last 2.4 Ma). Raymo et al. (1998) suggest that dramatic climate events were not confined to the last few large glacial-interglacial cycles. Looking at a relatively high-resolution ocean core in the North Atlantic, Raymo et al. found that all through the last 1.5 Myr, large sudden climate events (resembling Heinrich events and interstadials) occurred during the cooler parts of the climate cycles. This further suggests that dramatic instability is a 'normal' part of the Earth's climate system during this phase in its history, and that it is not merely confined to the extreme glacial-interglacial oscillations that have only operated for the last 900,000 years.

II.2.d. Sudden climate jumps from the more recent past

Different sources seem to suggest differing speeds and intensities for Holocene climate events. The Little Ice Age began in late Medieval times and played a role in extinguishing Norse colonies on Greenland (e.g., Barlow et al., 1997). It ended at about 1650 AD (Bradley and Jones, 1992), and may have been the most rapid and largest change in polar circulation during the Holocene according to chemical indicators of windblown sea salt in the GISP2 ice core (O'Brien et al., 1997). The event even shows up in the marine

record in the northern Atlantic (Keigwin, 1996), but may have been of less importance in other regions (Bradley and Jones, 1992). If so, this would seem to imply that an event which was clearly intense in some regions (such as the dry phase around 8,200 years ago across so many low and mid-latitude regions), was not relatively so important in other regions (such as near to the poles). The Little Ice Age was thus just another climate oscillation (fairly small by comparison with many of the events recorded in ice cores and sediment records) which gave cooler conditions over the lands around the North Atlantic between about 700 and 200 years ago.

Additional, smaller changes are observed in the detailed Greenland ice cap record, but it is important to note that not all the rapid changes observed in the Greenland ice cap correspond to large climate changes elsewhere. For example, a warming of 4 deg.C per decade was observed in an ice core from northern Greenland for the 1920's (Dansgaard et al. 1989), but this corresponded to a global shift of 0.5 deg.C or less. There is some evidence that this event may have been widespread (Thompson et al., 1993), but by no means has it been demonstrated to have been global. For this reason it is always desirable to have sources of evidence from other regions before invoking a broad, dramatic climate shift. What this relatively recent climate shift does suggest though, is that the climate system tends to undergo most of its changes in sudden jumps, even if those changes are relatively small against the background of those seen during the Quaternary. This is further evidence that if and when the next climate shift occurs, it will not be a gradual century-on-century change but rather a sudden step-function that will begin suddenly and occur over a decade or two.

III. The mechanisms behind sudden climate transitions.

It is still unclear how the climate on a regional or even global scale can change as rapidly as present evidence suggests. It appears that the climate system is more delicately balanced than had previously been thought, linked by a cascade of powerful mechanisms that can amplify a small initial change into a much larger shift in temperature and aridity (e.g., Rind and Overpeck, 1993). At present, the thinking of climatologists tends to emphasize several key components:

III.1. North Atlantic circulation as a trigger or an amplifier in rapid climate changes.

The circulation of the north Atlantic Ocean probably plays a major role in either triggering or amplifying rapid climate changes in the historical and recent geological record (Broecker 1995, Keigwin et al., 1994, Jones et al., 1996; Rahmstorf et al., 1996). The North Atlantic has a peculiar circulation pattern: the north-east trending Gulf Stream carries warm and relatively salty surface water from the Gulf of Mexico up to the seas between Greenland, Iceland and Norway. Upon reaching these regions, the surface waters cools off and (with the combination of being cooler and relatively salty because it mixes with mid-depth overflow water from the Mediterranean) becomes dense enough to sink into the deep ocean. The 'pull' exerted by this dense sinking water is thought to help maintain the strength of the warm Gulf Stream, ensuring a current of warm tropical water into the north Atlantic that sends mild air masses across to the European continent (e.g., Rahmstorf et al., 1996; Schmitz, 1995) (Fig. 3).

If the sinking process in the north Atlantic were to diminish or cease, the weakening of the warm Gulf Stream would mean that Europe had colder winters (e.g., Broecker, 1995). However, the Gulf Stream does not give markedly warmer summers in Europe - more the opposite in fact - so a shutting off of the mild Gulf Stream air masses does not in itself explain why summers also became colder during sudden cooling events (and why ice masses started to build up on land due to winter snows failing to melt during summer). In the North Atlantic itself, sea ice would form more readily in the cooler winter waters due to a shut-off of the Gulf Stream, and for a greater part of the year the ice would form a continuous lid over the north Atlantic. A lid of sea ice over the North Atlantic would last for a greater proportion of the year; this would reflect back solar heat, leading to cooler summers on the adjacent landmass as well as colder winters (e.g., Jones et al., 1996; Overpeck et al., 1997). With cooler summers, snow cover would last longer into the spring, further cooling the climate by reflecting back the sun's heat. The immediate result of all this would be a European and west Siberian climate that was substantially colder, and substantially drier because the air that reached Europe would carry less moisture, having come from a cold sea ice surface rather than the warm Gulf Stream waters.

After an initial rapid cooling event, the colder summers would also tend to allow the snow to build up year-on-year into a Scandinavian ice sheet, and as the ice built up it would reflect more of the Sun's heat, further cooling the land surface, and giving a massive high pressure zone that would be even more effective at diverting Gulf Stream air and moisture away from the mid-latititudes of Europe. This would reinforce a much colder regional climate.

The trigger for a sudden 'switching off' or a strong decrease in deep water formation in the North Atlantic must be found in a decrease in density of surface waters in the areas of sinking in the northern Atlantic Ocean. Such a decrease in density would result from changes in salinity (addition of fresh water from rivers, precipitation, or melt water), and/or increased temperatures (Dickson et al., 1988; Rahmstorf et al., 1996). For example, an exceptionally wet year on the landmasses which have rivers draining into the Arctic sea (Siberia, Canada, Alaska) would lead to such a decreased density. Ocean circulation modelling studies suggest that a relatively small increase in freshwater flux (called 'polar halocline catastrophe') to the Arctic Sea could cause deep water production in the north Atlantic to cease (e.g., Mikolajewicz and Maier-Reimer, 1994; Rahmstorf, 1994; Rahmstorf et al., 1996; 141).

During glacial phases, the trigger for a shut-off or a decrease in deep water formation could be the sudden emptying into the northern seas of a lake formed along the edge of a large ice sheet on land (for instance, the very large ice-dammed lake that existed in western Siberia), or a diversion of a meltwater stream from the North American Laurentide ice sheet through the Gulf of St. Lawrence , as seems to have occurred as part of the trigger for the Younger Dryas cold event (e.g., Kennett, 1990; Berger and Jansen, 1995). A pulse of fresh water would dilute the dense, salty GulfStream and float on top, forming a temporary lid that stopped the sinking of water that helps drive the Gulf Stream. The Gulf Stream could weaken and its northern end (the North Atlantic Drift) could switch off altogether, breaking the 'conveyer belt' and allowing an extensive sea ice cap to form across the North Atlantic, preventing the ocean current from starting up again at its previous strength. Theoretically, the whole process could occur very rapidly, in the space of just a few decades or even several years. The result could be a very sudden climate change to colder conditions, as has happened many times in the area around the North Atlantic during the last 100,000 years.

The sudden switch could also occur in the opposite direction, for example if warmer summers caused the sea ice to melt back to a critical point where the sea ice lid vanished and the Gulf Stream was able to start up again. Indeed, following an initial cooling event the evaporation of water vapour in the tropical Atlantic could result in an 'oscillator' whereby the salinity of Atlantic Ocean surface water (unable to sink into the north Atlantic because of the lid of sea ice) built up to a point where strong sinking began to occur anyway at the edges of the sea ice zone. The onset of sinking could result in a renewed northward flux of warm water and air to the north Atlantic, giving a sudden switch to warmer climates, as is observed many times within the record of the last 130,000 years or so.

The process of switching off or greatly dimishing the flow of the Gulfstream would not only affect Europe. Antarctica would be even colder than it is now, because much of the heat that it does receive now ultimately comes from Gulf Stream water that sinks in the north Atlantic, travels in a sort of river down the western side of the deep Atlantic Basin and then rat least partially esurfaces just off the bays of the Antarctic coastline (e.g., Schmitz, 1995). Even though this water is only a few degrees above freezing when it reaches the surface, this water is much warmer than the adjacent Antarctic continent, helping to melt back some of the sea ice that forms around Antarctica in the ice-free regions called polynyas. The effect of switching off the deepwater heat source would be cooler air and a greater sea ice extent around Antarctica, reflecting more sunlight and further cooling the region. However, the north Atlantic deep water takes several hundred years to travel from its place of origin to the Antarctic coast, so it could only produce an effect a few centuries after the change occurred in the North. It is not known what delay was present in the various climate changes that occurred between the north Atlantic region and Antarctica, but it is generally thought that other (relatively indirect) climate mechanisms, such as greenhouse gases in the atmosphere, linked these two far-flung regions and sometimes produced closely synchronised changes (i.e. within a few centuries of one another).

Although the end of the Last Glacial and various other sudden climate events such as Heinrich events do show up in the Antarctic ice record, not all large changes show such a closely linked occurrence and timing

around the world. For example there is no clear trace of the Younger Dryas in the Vostok ice core from Antarctica (Chapellaz et al. 1993; Broecker, 1998), and the warming at the start of the Eemian also does not seem to particularly closely linked to the timing of the warming which took place in the northern latitudes (Sowers et al. 1993).

During the colder glacial phases, deep water formation in the present areas between Greenland, Iceland and Norway would have ceased or dimished due to a thick cap of sea ice (though there is evidence it occasionally opened up to let Gulf Stream water through to the sea between Iceland and Norway, this did not result in much deepwater formation and so the pull and the northward heat flux seems to have been small). Instead, during the most intense cold phases the deepwater formation area seems to have moved to the south of the British Isles, at the edge of the extended sea ice zone (e.g., , Imbrie et al. 1992, Duplessey et al. 1984; Maslin et al., 1997). Even here, deep water formation seems to have been weaker than at present, producing relatively small quantities which penetrated to mid-depths rather than to the deepest ocean basins. This was probably at least partly because the whole surface of the Atlantic Ocean (even the tropics) was cooler; with less evaporation from its surface, even the water that did reach northwards was less briney (and thus less dense), so less able to sink when it reached the cold edge of the sea ice zone. An initial slowdown of north Atlantic circulation may have been the initial trigger for a set of amplifying factors (see below) that rapidly led to a cooling of the tropical Atlantic, reinforcing the sluggish state of the glacial-age Gulf Stream.

The idea of Gulf Stream slowdowns as a mechanism in climate change is not merely theoretical. There is actually evidence from the study of ocean sediments that deepwater formation in the north Atlantic was diminished during the sudden cold Heinrich events and other colder phases of the last 130,000 years, including the Younger Dryas phase (e.g., Fairbanks, 1989; Kennett, 1990; Maslin, 199x). The same appears to have been true further back in time to 1.5 Myr ago (Raymo et al. 1998). The process also 'switched on' rapidly at times when climates suddenly warmed around the north Atlantic Basin, such as at the beginning of interstadials or the beginning of the present interglacial (Ramussen et al. 1997). Decreasing deep water formation occurred at times when the climate was cooling towards the end of an interstadial, and it diminished suddenly with the final cooling event that marked the end of the interstadial (Ramussen et al., 1997), and over a period of less than 300 years at the beginning of the Younger Dryas (e.g., Berger and Jansen, 1995).

The Intra-Eemian cooling event - if it really did happen - is perhaps one of the most alarming of the changes observed from recent climate history, because it occurred in the midst of an interglacial not too dissimilar from our own. There are strong signs that a rapid cut-off in NADW formation was associated with this climate change (Maslin and Tzedakis, 1996). This event near 122 ka seems to have coincided with intense cooling and freshening in the seas west of Ireland and in the northern Norwegian Sea (Cortijo et al., 1994; Fronval and Jansen, 1996). Today any similar reduction in the salinity of the sea surface in these two areas would be enough to result in a reduction of deep water formation in the Nordic Seas (e.g. Bryan, 1986; Dickson et al., 1990; Seidov and Maslin, 1996). This reduction could explain why Europe (and perhaps other areas) cooled so dramatically (Maslin et al. 1998).

What might have caused this freshening of the North Atlantic seas? The fresh water could have come from melting icebergs, from increased rain and snowfall in the region, or a change in surface currents that brought in fresher water from another part of the ocean. There is no evidence of an increased supply of melting icebergs (Keigwin et al., 1994; McManus et al., 1994) during this phase. Instead, one or more of the following factors seems a more likely cause: 1) greater incursion of relatively 'fresh' North Pacific water through the Bering Strait and around the Arctic Seas to the North Atlantic during times of raised sea level during the Eemian (Shaffer and Bendtsen, 1994), 2) enhanced precipitation over the North Atlantic and seas between Iceland and Norway, due to the greater amount of solar heat reaching the northern latitudes during the summer at that time (due to the Milankovitch rhythms in the Earth's orbit; see below), and 3) the closure of the Northern Baltic sea link to the North Atlantic ocean, which would ensure that more fresh water ended up in the area (van Andel and Tzedakis, 1996).

We do not yet know how widespread the event was, or what caused it, but its occurrence at least points to NADW formation being a 'weak point' in the interglacial climate system (e.g., Broecker, 1997), with potential to affect climate rapidly in the North Atlantic region and perhaps elsewhere. This is rather

worrying, for we are presently in the midst of an interglacial phase not greatly unlike the Eemian.

Broader changes in temperature and rainfall over much of the world are thought likely to have occured as a result of a switching on or off of the north Atlantic circulation (Jones et al., 1996; Rind and Overpeck, 1993), and these changes would result in amplification by the feedback mechanisms suggested below. As evidence of such a broader link to global climate, over recent years changes in the monsoon-belt climates of Africa and Asia have also been observed to occur in association with decadal-scale phases of weaker north Atlantic circulation (e.g., Hurrel, 1995, 1996). By extrapolation, it is generally thought that bigger changes in the north Atlantic circulation would result in correspondingly larger changes in climates in the monsoon belts and in other parts of the world.

In addition to this relatively direct effect of deepwater on North Atlantic and Antarctic climate, other subtle effects on global climate would be expected to result from a sudden change in north Atlantic circulation, or indeed they may themselves trigger a change in the north Atlantic circulation by their effects on atmospheric processes. These include the interaction with global carbon dioxide concentrations, dust content and surface reflectivity (albedo).

II.2 Carbon dioxide and methane concentration as a feedback in sudden changes.

Analysis of bubbles in ice cores shows that at the peak of glacial phases, CO2 was about 30% lower than during interglacial conditions (e.g., Jouzel et al., 1993). We do not at present know whether the lower glacial CO2 levels were a cause or merely an effect of the ice ages. Lower CO2 level might result from changes in plankton productivity (e.g., Lyle et al., 1988; see also discussion in Thomas et al., 1995), drawing more carbon down out of the atmosphere once climate began to cool. Terrestrial biomass reservoirs if anything seem to move in the opposite direction, releasing carbon to the atmosphere with the onset of glaciation (Adams & Faure 1998). The lower carbon dioxide concentrations resulting from greater ocean carbon storage would cool the atmosphere, and allow more snow and ice to accumulate on land. Relatively rapid changes in climate, occurring over a few thousand years, could have resulted from changes in the atmospheric CO2 concentration (e.g., Broecker, 1997). The actual importance of carbon dioxide in terms of the climate system is unknown, though computer climate simulations tend to suggest that it directly cooled the world by less than 1 deg.C on average, but due to amplification of this change by various factors within the climate system such as the water vapour content, the resulting change in global climate could have been more than 2 deg.C (e.g., Houghton et al., 1995). In addition, warming as a result of increased atmospheric CO2 levels is greater at higher latitudes, possibly inducing snow and ice melt.

A problem with invoking atmospheric carbon dioxide levels as a causal factor in sudden climate changes is that they seems to have varied too slowly, following on the timescale of millennia what often occurred on the timescale of decades- but the resolution of our records may not be good enough to resolve this question now. Methane, a less important greenhouse gas, was also 50% lower during glacial phases (e.g., Sowers et al., 1993), probably due to reduced biological activity on the colder, drier land surfaces (Meeker et al., 1997). However, it does seem to have increased rapidly in concentration in association with changes in climate, reaching its normal Holocene levels in around 150 years or less during the global climate warming at the end of the Younger Dryas, around 11,500 years ago (Taylor et al. 1997).

Such sudden rises in atmospheric methane concentration were probably not important in affecting climate; the warming effect of a 50% change in methane would have been much less than an equivalent change in CO2, because methane is at such a low overall concentration in the atmosphere. It has been suggested that another mechanism, involving sudden and short-lived releases of massive amounts of methane from the ocean floors, could sometimes have resulted in rapid warming phases that do not leave any trace in terms of raised methane levels in the ice core data, where the trapped gas bubbles generally only indicate methane concentrations at a time resolution of centuries rather than the few years or decades that such a 'methane pulse' might last for. However, ice cores from areas where the ice sheet built up particularly rapidly (Chapellaz et al. 1993, Taylor et al. 1997) show detailed time resolution of the record of methane concentration in the atmosphere, and fail to show evidence of sudden 'bursts' of methane. However, the idea does remain a possibility, to at least some biogeochemists (Brook et al. 1996).

III.3 Surface reflectivity (albedo) of ice, snow and vegetation.

The intensely white surface of sea ice and snow reflects back much of the sun's heat, hence keeping the surface cool. Presently, about a third of the heat received from the sun is reflected back into space, and changes in this proportion thus have the potential to strongly influence global climate (e.g., Crowley and North, 1991). In general the ice cover on the sea, and the snow cover on the land, have the potential to set off rapid climate changes because they can either appear or disappear rapidly given the right circumstances. Ice sheets are more permanent objects which, whilst they reflect a large proportion of the sunlight that falls upon them, take hundreds of years to melt or build up because of their sheer size. When present, sea ice or snow can have a major effect in cooling regional and global climates, but with a slight change in conditions (e.g. just a slightly warmer summer) they will each disappear rapidly, giving a much greater warming effect because sunlight is now absorbed by the much darker sea or land cover underneath. In an unusually cold year, the opposite could happen, with snow staying on the ground throughout the summer, itself resulting in a cooler summer climate. A runaway change in snow or sea ice (positive feedback) could thus be an important amplifier or trigger for a major change in global temperature. It is possible that by slow changes over millennia or centuries, the climate could be brought to a break point involving a runaway change in snow and ice reflectivity over a few decades. These slow background changes might include variations in the earth's orbit (affecting summer sunlight intensity), or gradual changes in carbon dioxide concentration, or in the northern forest cover which affects the amount of snow that is exposed to sunlight.

It is possible that the relatively long-lived ice sheets might occasionally help bring about very rapid changes in climate, by rapidly 'surging' outwards into the sea and giving rise to large numbers of icebergs that would reflect back the sun's heat and rapidly cool the climate (e.g., MacAyeal, 1993a, b). The intensely cold Heinrich events that punctuated the last ice age were initially thought to be caused by sudden slippage of the Laurentide ice sheet that covered most of Canada. It appears, however, that all the separate ice sheets around the north Atlantic surged outwards simultaneously, and that their outwards movement probably thus represents a secondary response to an initial climate cooling (e.g. a change in the deepwater formation system in the north Atlantic) rather than the initial trigger (Bond et al., 1997.). This does not mean that ice surges and ice bergs were irrelevant in the extreme cold of Heinrich events; by their albedo effects they may have helped to intensify and temporarily stabilize a cooling event that would have occurred anyway. However, this amplification may have occurred decades or centuries after the initial 'step function' event associated with the rapid cooling.

Another, possibly neglected, factor in rapid regional or global climate changes may be the shifts in the albedo of the land surface that result from changes in vegetation or algal cover on desert and polar desert surfaces. An initial spreading of dark-coloured soil surface algae or lichens following a particularly warm or moist year might provide a 'kick' to the climate system by absorbing more sunlight and thus warming the climate, and also reducing the dust flux from the soil surface to the atmosphere (see below). Larger vascular plants and mosses might have the same effect on the timescale of years or decades. The detailed analysis of the ending of the Younger Dryas by Taylor et al. 1997, suggests that warming occurred around 20 years earlier in lower and mid latitudes, perhaps due to some initial change in vegetation or snow cover affecting land surface albedo. Some of the earlier climate warming events during the last 130,000 years show similar signs of changes in dust flux followed by changes in high-latitude temperature (Raymo pers. comm. 1997).

III.4 Water vapour as a feedback in sudden changes.

Water vapour is a more important greenhouse gas than carbon dioxide, and as its atmospheric concentration can vary rapidly, it could have been a major trigger or amplifier in many sudden climate changes. For example, a change in sea ice extent or in carbon dioxide, would be expected to affect the flux of water vapour into the atmosphere from the oceans, possibly amplifying climate changes. Large, rapid changes in vegetation cover might also have added to these changes in water vapour flux to the atmosphere. In a recent (1997) lecture presentation and an article in Atlantic Monthly (December 1997) the geologist W.S. Broecker suggested that water vapour may act as a global 'messenger', co-ordinating rapid climate changes, many of which seem to have occurred all around the world fairly simultaneously, or in close succession. Broecker notes the evidence for large changes in the water vapour content of the atmosphere in terms of changes in the 18O content of tropical high Andean ice cores (Thompson et al. 1995).

III.5. Dust and particulates as a feedback in sudden changes.

Particles of mineral dust, plus the aerosols formed from fires and from chemicals evaporating out of vegetation and the oceans, may also be a major feedback in co-ordinating and amplifying sudden large climate fluctuations. Ice cores from Greenland (Taylor et al. 1997; Mayewski et al., 1997), Antarctica (Jouzel et al., 1996) and tropical mountain glaciers (Thompson et al., 1987; 1995) show greater concentrations of mineral dust during colder phases. This suggests that there was more dust around in the world's atmosphere during cold periods that during warm phases. It seems that the atmospheric content of dust and sulphate particles changed very rapidly, over just a few decades, during sudden climate transitions in the Greenland ice core record (Taylor et al. 1997). The drier and colder the world gets, the more desert there is and the higher the wind speeds, sending more desert dust into the atmosphere where it may reinforce the cold and dryness by forming stable 'inversion' layers that block sunlight and prevent rain-giving convective processes. A run of wet years in the monsoon belt could trigger rapid revegetation of desert surfaces by vascular plants or algae, and a sudden decrease in the amount of dust blown into the atmosphere. Less dust could help make conditions still warmer and wetter, pushing the climate system rapidly in a particular direction (though dust and other particles might actually tend to warm the surface if they blow over lighter-coloured areas covered by snow or ice; Overpeck et al. 1997).

It has even been proposed that variations in the influx of dust from outer space (Interplanetary Dust Particles, IDPs) could have played a role in triggering the large-scale glacial/interglacial alternations at 100 kyr periodicities (Muller and MacDonald, 1997; Farley and Paterson, 1995). It was proposed that the accretion rate of IPDs might be linked to the varying inclination of the Earth's orbit with respect to the inavriable plane of the solar system. It now appears, however, that this process is not likely to have been of influence on ice ages on Earth (Kortenkamp and Dermott, 1998).

Haze production is a poorly understood but potentially very significant factor in triggering or amplifying sudden climate changes. Given what is known of the present-day patterns of emission of haze-producing compounds from land vegetation, decade-to-century timescale changes in vegetation distribution and activity could have resulted in rapid changes in global haze production (Adams et al. submitted manuscript), although because this factor eludes the sedimentological record, it may be impossible to test it with observations from the past.

III.6. Seasonal sunlight intensity as a background to sudden changes.

A major background factor in pacing climate switches on timescales of tens of thousands of years seems to have been the set of 'Milankovitch' rhythms in seasonal sunlight distribution or insolation (Imbrie and Imbrie, 1992; Imbrie et al., 1992, 1993). Although the insolation values change gradually over thousands of years, they may take the earth's climate to a 'break point' at which other factors will begin to amplify change into a sudden transition.

In one of the Milankovitch rhythms, the shape of the earth's orbit shifts from more elliptical to more nearly circular at periodicities of about 100 kyr, which alters the total amount of solar radiation received on Earth. In another the degree of tilt of the earth's axis changes (periodicty of 42 kyr), and in the third the timing of the seasons changes relative to the earth's elliptical track nearer and further from the sun (periodicities of 19 and 23 kyr). These latter two rhythms alter the relative amount of solar radiation reaching the Earth's Northern and Southern Hemispheres during summer and winter. Times when summer sunlight in the Northern Hemisphere is strong (but when the winter sunlight is correspondingly weak) tend to be the times when the rapid global transition from glacial to interglacial conditions occurs.

These big glacial-interglacial transitions roughly follow the 100,000-year timescale during the last 900 kyr, when the three different rhythms (and possibly the poorly understood factors such as the internal structure of ice-sheets; MacAyeal, 1993a, b) line up to give a big increase in northern summer warmth. However the lesser individual rhythms can also be detected in the temperature record on the 19,000 and 42,000-year timescales, and in fact the timing of interglacial onset tends to more closely follow multiples of the 19,000 year cycle than an exact correspondence to the 100,000 year cycle (Imbrie et al. 1992, 1993). This is thought to be due to the effects of summer temperatures on various of the factors mentioned above; for

example, it ensures melting back of snow and sea ice in summer, helping the earth to absorb more solar radiation and thus to heat up further. It is generally accepted that the effect of changes in heat budget as a result of the Milankovitch variations by themselves are not enough to bring about the large, rapid changes in climate that follow these rhythms in seasonal sunlight, and some set of positive feedback factors - directly or indirectly linked to the seasonal insolation changes - must be involved to bring the earth out of glacial and into interglacial conditions.

It is important to note, however, that most of the very rapid climate transitions during the last 100,000 years do not show any clear association in timing with the background Milankovitch rhythms, especially the fluctuations at periodicities below 19 kyr. In these cases their ultimate trigger must lie in other factors, probably a combination of many processes that sometimes line up to set the climate system on a runaway course in either the direction of cooling or warming.

IV. Could dramatic decade-timescale climate transitions occur in the near future?

From present understanding of the record of the last 150,000 years, at least a few large climate changes certainly occurred on the timescale of individual human lifetimes, the most well-studied and well-established of these being the ending of the Younger Dryas, and various Holocene climate shifts. Many other substantial shifts in climate took at most a few centuries, and they too may have occurred over a few decades. The high time resolution in the climate record, however, is either not available, or records have not yet been studied in enough detail. Some very interesting new data sets may be expected to become available within a few years, as a result of drilling by the Ocean Drilling Program in the Saanich Inlet (western Canada; Leg 169A) and in the northern Atlantic (Leg 172). It will take time before the meticulous work of logging year-by-year changes in long ice cores and lake records can give a relatively complete picture of when, and exactly how quickly, rapid climate changes occurred. There are many 'suspected' decade-timescale climate changes from the past (just as the Younger Dryas was until recently a 'suspected' but distinctly unproven decadal-scale climate shift), but very few 'proven' ones. Greater knowledge of how frequently such sudden events have occurred, and under what general circumstances, is required before a greater understanding can be reached.

It is difficult to say what the risks are of a sudden switch in global or North Atlantic region climate, because the mechanisms behind all past climate changes (sudden or otherwise) are incompletely understood. They appear to be real, however, and relatively small-scale changes in North Atlantic salinity have been observed and studied in the last few decades (Dickson et al., 1988). Fluctuations in surface water characteristics and precipitation patterns in that region vary on decadal time scales with variations in the strength of high-pressure areas over the Azores and Iceland (the North Atlantic Oscillation; Hurrel, 1995, 1996), providing an observed apparent link between salinity and climate fluctuations. The fear is that relatively small anthropogenic changes in high-latitude temperature as a result of increased concentrations of greenhouse gases might switch North Atlantic circulation and alter the course of the Gulfstream during such natural fluctuations.

Not even knowing how often decade-timescale changes occurred in the recent geological past, we are handicapped in trying to find mechanisms which might explain them and be used for forecasting future events. Even if one knew everything there was to know about past climate mechanisms, it is likely that we would still not be able to forecast such events confidently into the future. This is because the system will have been influenced by probabilistic processes (due to the chaotic nature of the ocean-climate system, with runaway changes coming from miniscule differences in initial conditions; e.g. Crowley and North, 1991), so it is not justifiable to talk in terms of what 'definitely' will or will not happen in the future - even though the public and policymakers are looking for certainities. All that one can reasonably do is set out what the current understanding is, acknowledging that this understanding is limited and may turn out to be wrong in certain key respects, and then talk in terms of probablities of particular events occurring.

At the outset, there is the possibility that most of the climate instability seen in the recent geological past is not relevant to our immediate future, because it represents a different system; a 'glacial' state, almost certainly charcaterized by a different pattern of deep-sea circulation (e.g., Rahmstorf et al., 1996). Most of the rapid climate transitions during the last 150,000 years seem to have occurred against the background of a world with a larger northern ice sheet extent than at present, perhaps indicating that in this glacial mode

the climate is predisposed to be more unstable than in our present interglacial state (e.g., Rind and Chandler, 1991). Even the sudden and widespread early-to-mid Holocene arid event (8,200 y.a.) occurred at a time when large parts of the Laurentide ice sheet remained unmelted over Canada.

However, there were at least some rapid climate transitions which occurred when ice sheet extent was no greater than at present, such as the apparently widespread late Holocene cool/arid events at 8200 yrs BP, at around 3,800 yrs BP, and another cool event around 2,600 yrs BP. (although the time taken for onset of these later Holocene changes in regional and global climates does not yet seem to have been determined).

Various large full-interglacial climate changes during the Holocene and certain earlier interglacials (e.g. the Eemian and the Holstein Interglacials in Europe; Winograd et al. 1997) that show up in the Greenland ice cap also seem to correlate with genuinely large climate shifts in Europe and elsewhere, taking conditions from temperate to boreal or even sub-arctic. Whether they occurred over decades, centuries or thousands of years, they offer a worrying analogue for what might happen if greenhouse gas emissions continue unchecked. Judging by its past behaviour under both glacial (e.g. the ending of the Younger Dryas) and interglacial conditions (e.g. the various Holocene climate oscillations leading up to the 20th century; Alley et al., 1997), climate has a tendency to remain quite stable for most of the time and then suddenly 'flip'; at least sometimes over just a few decades, due to the influence of the various triggering and feedback mechanisms discussed above. Such observations suggest that even without anthropogenic climate modification there is always an axe hanging over our head, in the form of random very large-scale changes in the natural climate system; a possibility that policy makers should perhaps bear in mind with contingency plans and international treaties designed to cope with sudden famines on a greater scale than any experienced in written history. By starting to disturb the system, humans may simply be increasing the likelihood of sudden events which could always occur.

Another source of evidence seems to underline the potential importance of sudden climate changes in the coming centuries and millennia: computer modelling studies of the (still incompletely understood) north Atlantic deepwater formation system suggest that it is indeed sensitive to quite small changes in freshwater runoff from the adjacent continents, whether from river fluxes or meltwater from ice caps (Rahmstorf et al., 1996). Some scenarios in which atmospheric carbon dioxide levels are allowed to rise to several times higher than at present result in increased runoff from rivers entering the Arctic Basin, and a rapid weakening of the Gulf Stream, resulting in colder conditions (especially in winter) across much of Europe. Just doubling the amount of carbon dioxide in the atmosphere could be enough to set off such a change (Broecker, 1997). Whilst these are only preliminary models, and thus subject to revision as more work is done, they do seem to point in the same direction as the ancient climate record in suggesting that sudden shutdowns or intensification of the Gulf Stream circulation might occur under full interglacial conditions, and be brought on by the disturbance caused by rising greenhouse gas levels. To paraphrase W.S. Broecker; 'Climate is an ill-tempered beast, and we are poking it with sticks'.

Acknowledgements. Many of our colleagues have offered helpful advice on the Quaternary record of climate history. Jonathan Adams is supported by a Eugene P. Wigner fellowship from the Lockheed-Martin Corporation, at the Environmental Sciences Division of Oak Ridge National Laboratory.

Adams J.M. & Faure H. 1998: A new estimate of changing carbon storage storage on land since the last glacial maximum, based on global land ecosystem reconstruction. in press, Global and Planetary Change.

Adkins J. F., Boyle E. A., Keigwin, L., and Cortijo E., 1997: Variability of the North Atlantic thermohaline circulation during the last interglacial period. Nature 390 154-156.

Alley, R. B., Mayewski, P. A., Sowers, T., Stuiver, M., Taylor, K. C., and Clark, P. U., 1997: Holocene climatic instability: a prominent, widespread event 8200 yr ago. Geology, 25, 483-486.

Andrews, J. T., Erlenkeuser, H., Tedesco, K., Aksu, A. E., and Jull, A. J. T., 1994: Late Quaternary (Stage 2

and 3) Meltwater and Heinrich Events, Northwest Labrador Sea, Quaternary Research 41 26-34.

Bahn, P. G., and Everett, K., 1993: Iceman in the cold light of day. Nature, 362, 11-12.

Barlow, L. K., Sadler, J. P., Ogilvie, A. E. J., Buckland, P. C., Amorosi, T., Ingimundarson, J. H., Skidmore, P., Dugmore, A. J., and McGovern, T, H., 1997: Interdisciplinary investigations of the end of the Norse Western Settlement in Greenland, The Holocene, 7, 489-499.

Baroni, C., and Orombelli, G., 1996: The Alpine 'Iceman' and Holocene Climatic Change. Quaternary Research, 46, 78-83. Behl, R. J., and Kennett, J. P., 1996, Brief interstadial events in the Santa Barbara basin, NE Pacific during the past 60 kyr. Nature, 379, 243-246.

Berger , W., Bickert, T., Jansen, E., Wefer, G., and Yasuda, M., 1993: The Central Mystery of the Quaternary Ice Age, Oceanus, 36, 53-56

Bjorcke, S., Kromer, B., Johnsen, S., Bennike, O., Hammarlund, D., REF?

Berger, W. H., and Jansen, E., 1995: Younger Dryas episode: ice collapse and super-fjord heat pum IN: The Younger Dryas, S. R. Troelstra, J. E. van Hinte, and G. M. Ganssen, eds., Koninklijke Nederlands Akademie van Wetenschappen, Afdeling Natuurkunde, Eerste Reeks, 44, 61-105.

Lemdahl, G., Possnert, G., Rasmussen, T. L., Wohlfarth, B., Hammer, C. U., and Spurk, M., 1996: Synchronized terrestria-atmospheric deglaical records around the North Atlantic. Science, 274, 1155-1160.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManus, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G., and Ivy, S., 1992: Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period. Nature, 360, 245-249.

Bond, G. C., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J., and Bonani, G., 1993: Correlations between climate records from North Atlantic sediments and Greenland ice. Nature, 365, 143-147.

Bond, G. C., and Lotti, R., 1995: Iceberg Discharges into the North Atlantic on Millennial Time Scales During the last deglaication. Science, 267, 1005-1010.

Bond, G., Showers, W., Cheseby, M., Lotti, R., Almasi, P., deMenocal, P., Priore, P., Cullen H., Hajdas I. & Bonani G., 1997: A pervasive millenial-scale cycle in North Atlantic Holocene and Glacial climates. Science 278 1257-1265.

Boulton,G. S., 1993: Two cores are better than one, Nature, 366, 507-508

Bradley, R. S., and Jones, P., eds., 1988: Climate since AD 1500. Chapman and Hall, NY, 678

Bradley, R. S., and Jones, P. 1992: The Little Ice Age. The Holocene 3, 367-376

Bradley, R. S., and Jones, P., eds., 1992a: Climate since AD 1500. Chapman and Hall, NY, 678

Broecker W.S. 1995: 'The glacial world according to Wally'. Eldigo press Broecker, W. S., Bond, G., Klas, M., Clark, E., and McManus, J., 1992. Origin of the North Atlantic's Heinrich Events. Climate Dynamics, 6, 265-273

Broecker W.S., 1997: "Will Our Ride into the Greenhouse Future be a Smooth One?" GSA Today, 7(5):1-7

Broecker, W. S., 1997: Thermohaline circulation, the Achilles heel of our climate system: will man-made CO_2 upset the current balance? Science,. 278, 1582-1588.

Broecker, W., 1998: Paleocean circulation during the last deglaciation: a bipolar seesaw? Paleoceanography, 13, 119-121.

Brook, E. J., Sowers, T., and Orchardo, J., 1996: Rapid Variations in atmosheric methane concentration during the past 110,000 years. Science 273, 1087-1091

Bryan, K., 1986: Poleward buoyancy transport in the ocean and mesocscale eddies, Journal of Physical Oceanography, 16, 927-933.

Campbell, I. D., Campbell, C., Apps, M. J., Rutter, N. W., and Bush, A. B. G., 1998: Late Holocene ~1500 yr periodicities and their implications, Geology, 26, 471-473.

Chappellaz J.M., Blunier T., Ratnaud D., Barnola J.M., Schwander J. & Stauffer B., 1993: Synchronous changes in atmospheric CH4 and Greenland climate between 40 and 8 kyr BP. Nature 366 443-445.

Cortijo E., Duplessy, J.-C., Labaeyrie, L., Leclaire, H., Duprat, J., and van Weering, T. C. E., 1994: Eemian cooling in the Norwegian Sea and North Atlantic Ocean preceding continental ice-sheet growth. Nature, 372, 446-449.

Crowley T. J. and North G. R., 1991: Palaeoclimatology. Oxford University Press, Oxford

Curtis, J. H., Hodell, D. A., and Brnner, M., 1996: Climate variability on the Yucatan Peninsula (Mexico) during the past 3500 years, and implications for Maya cultural evolution. Quatearnary Research, 46, 47-47.

Dansgaard, W., White, J. W. C., and Johnsen, S. J., 1989: The abrupt termination of the Younger Dryas climate event. Nature, 339, 532-534.

Dansgaard, W., Johnson, S. J., Clausen, H. B., Dahl-Jensen, D., Gundenstrup, N. S., Hammer, C. U., Hvidberg, C. S., Steffensen, J.P., Sveinbjrnsdottir 1993: Evidence for general instability of past climate from a 250-kyr ice-core record. Nature 364, 218-220

de Beaulieu, J.-L., and Reille, M., 1989: The transition from temperate phases to stadials in the long upper Pleistocene sequence from les Echets (France), Palaeogeography, Palaeoclimatology, Palaeoecology, 72, 147-159

De menocal, P., 1995: Plio-Pleistocene African Climate. Science, 270, 53-59. Dickson, R. R., Meincke, J., Malmberg, S. A., and Lee, A. J., 1988, The "Great Salinity Anomaly" in the northern North Atlantic, 1968-1982. Progress in Oceanography, 20, 103-151

Dorale, J. A., Gonzalez, L. A., Reagan, M. K., Pickett, D. A., Murrell, M. T., and Baker, R. G., 1992: A high-resolution record of Holocene climate change in speleothem calcite from Cold Water Cave, Northeast Iowa. Science, 258, 1626-1630.

Duplessy, J. C., Shackleton, N. J., Matthews, R. K., Prell, W. L., Ruddiman, W. F., Caralp, M., and hendy, C., 1984: 13C record ofbenthic foraminifera in the last interglacial ocean: implications for the crabon cycle and the global deep water circulation. Quaternary Research 21, 225-243

Edwards, R. L., Cheng, H., Murreil, M. T., and Goldstein, S. J., 1997: Protactinium-231 dating of Carbonates by Thermal Ionization Mass Spectrometry: Implications for Quaternary Climate Change. Science 276, 782-786

Field M, Huntley B. & Muller H. 1994: Eemian climate fluctuations observed in a European pollen record. Nature 371, 779.

Farley, K. and D. B. Patterson, 1995: A 100 kyr periodicity in the flux of extraterrestrial 3He to the sea floor. Nature 378, 600-603.

Francois, R., and Bacon, M. P., 1993: Heinrich events in the North Atlantic: radiochemical evidence. Deep-Sea Research, XX, X-X Frenzel et al. 1992

Fronval, T., and Jansen, E., 1996: Rapid changes in ocean circulation and heat flux in theNordic seas druing the last interglacial period. Nature, 383, 806-810

Fuhrer, K., and Legrand, M., 1997: Continental biogenic species in Greenland Ice core Project core: Tracing back the biomass history of the North American continent. Journal of Geophysical Research, 102, 26,735-26,745

Grimm, E. C., Jacobson, G. L., Watts, W. A., Hansen, B. C. S., and Maasch, K., 1993: A 50,000 year record of climate oscillations from Florida and its temporal correlation with the Heinrich events. Science, 261, 198-200.

Grootes, P. M., Stuiver, M., White, J. W. C., Johnsen, S., and Jouzel, J., 1993. Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores, Nature, 366, 552- 554

Grousset, F. E., Labaeyrie, L., Sinko, J. A., Cremer, M., Bond, G. , Duprat, J., Cortijo, E., and Huon, S., 1993: Patterns of ice-rafted detritus in the glaical North Atlantic (40o-55oN), Paleoceanography, 8, 175-192.

Guiot , J. L., de Beaulieu, J. L., Chedaddi, R., David, F., Ponel, P., and Reille, M., 1993: The climate in Western Europe during the last glacial/interglacial cycle derived from pollen and insect remains, Palaeogeography, Palaeoclimatology, Palaeoecology, 103 73-93

Gwiazda, R. H., Hemming, S. R., and Broecker, W. S., 1996a: Tracking the sources of icebergs with lead isotopes: the provenance of ice-rafted debris in Heinrich layer 2. Paleoceanography 11, 77-93.

Gwiazda, R. H., Hemming, S. R., and Brocker, W. S., 1996b: Provenance of iceberg during Heinrich event 3 and the ocntrast to thier sources during other Heinrich events. Paleoceanography 11, 371-378.

Heinrich, H., 1988: Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years. Quaternary Research, 29, 142-152.

Hodell, D. A., Curtis, J. H., and Brenner, M., 1995: Possible role of climate in the collapse of Classic maya civilization. Nature, 375, 391-394.

Hodell, D. A., and Venz, K., 1992: Towards a high-resolution stable isotopic record of the Southern Ocean during the Pliocene-Pleistocene. AGU Antartic Research Series, 56, p 495-503.

Hurrell, J. W., 1995: Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. Science 269 676-679

Hurrell, J. W., 1996: Influence of variations in extratropical wintertime teleconnections on Northern Hemsphere temprature, geophys. Res. Lett., 23, 665-668.

Imbrie, J. and Imbie, K. P., 1992: Ice Ages: Solving the Mystery. Imbrie, J. Hays, J. D., Martinson, D. G., McIntyre, A., Mix, A. C., Morley, J. J., Pisias, N. G., Prell, W. L., Shackleton, N. J., 1984. The orbital theory of Pleistocene lcimate: support from a revised chronology of the marine 18O record. in: A. Berger, J. Imbrie, J. Hays, G. Kukla, B. Saltzman, eds., Milankovich and Climate, Reidel, Dordrecht, 269-306.

Imbrie, J., Boyle, E., Clemens, S., Duffy, A., Howard, W., Kukla, G., Kutzbach, J., Martinson, D., McIntyre, A., Mix, A., Molfino, B., Morley, J., Peterson, L., Pisias, N., Prell, W., Raymo, M., Sackleton, N., and Toggweiler, J., 1992: One the Structure and Origin of major Glaciation Cycles. 1. Linear responses to Milankovictch forcing. Paleoceanography, 7,701-738

Imbrie, J., Mix, A. C., and Martinson, D. G., 1993a: Milankovitch theory viewed from Devil's Hole. Nature, 363, 531-533.

Imbrie, J., Berger, A., Boyle, E. A., Clemens, S. C., Duffy, A., Howard, W. A., Kukla, G., Kutzbach, J., Martinson, D. G., McIntyre, A., Mix, A. C., Molfino, B., Morley, J. J., Peterson, L. C., Pisias, N. G., Prell, W.

G., Raymo, M. E., Shackleton, N. J., and Toggweiler, J. R., 1993b: On the Structure and Origin of Major Glaciation Cycles. 2. The 100,000 year cycle. Paleoceanography, 8, 699-735

Johnsen, S. J., Clausen, H. B., Dansgaard, W., Gundestrup, N. S., Hammer, C. U., and Tauber, H., 1995: The Eem satble isotope record along the GRIP ice core and its interpretation. Quaternary Research, 43, 117-124.

Jones, P. D., Bradley, R. S. and Jouzel, J., eds., 1996: Climatic variations and forcing mechanisms over the last 2,000 years. Springer

Jouzel, J., Barkov, N. I., Barnola, J. M., Bender, M., Chapellaz, J., Genthon, C., Kotlyakov, V. M., Lipenkov, V., Lorius, C., Petit, J. R., Raynaud, D., Raisbeck, G., Ritz, C., Sowers, T., Stievenard, M., Yiou, F., and Yiou, P., 1993, Extending the Vostok ice-coore record of palaeoclimate to the penultimate glacial period, Nature, 364, 407-411

Jouzel, J., Waelbroeck, C., Malaize, B., Bender, M., Petit, J. R., Stievenard, M., Barkov, N. I., Barnola, J. M., King, T., Kotlyakov, V. M., Lipenkov, V., Lorius, C., Raynaud, D., Ritz, C., and Sowers, T., 1996: Climatic interpretation of the recently extended Vostok ice records. Climate Dynamics 12, 513-521.

Keigwin, L., 1996: The Little Ice Age and Medieval Warm Period in the Sargasso Sea, Science, 274, 1504-1508.

Keigwin, L., Curry, W. B., Lehman, S. J., and Johnsen, S., 1994: The role of the deep ocean in North Atlantic climate change between 70 and 130 kyr ago, Nature 371 323-329

Kennet 1990 (ed.): , Special Section: The Younger Dryas Event, Paleoceanography, 5 891-1041

Kennett J.P. (1996): A review of polar climatic evolution during the Neogene, based on the marine sediment record. 49-64. In; Vrba E., Denton G.H., Partridge T.C. & Burckle L.H. Palaeoclimate and Evolution, with emphasis on human origins. Yale University Press, New Haven.

Kortenkamp, S. J., and Dermott, S. F., 1998: A 100,000 year Periodicity in the accretion rate of interplanetary dust. Science, 280, 874-876. Kukla, G., McManus, J., Rousseau, D.-D., and Chuine, I., 1997, How long and how stable was the last interglacial? Quaternary Science Reviews, 16, 605-612.

Labeyrie, L. D., Duplessy, J.-C., Duprat, J., Juillet-Leclerc, A., Moyes, J., Michel, E., Kallel, N., and Shackleton, N. J., 1992: Changes in the vetrical structure of the North Atlantic Ocean between glaical and modern times. Quaternary Science Reviews, 11, 401-413.

Lehman, S., 1996: True grit spells double trouble. Nature, 382, 25-26. Lowe J. and Walker, M. J. C., 1984. Reconstructing Quaternary Environments. Longman, NY

Lowell, T. V., Heusser, C. J., Andersen, B. G., Moreno, P. I., Hauser, A., Heusser, L. E., Schluchter, C., Marchant, D. R., and Denton, G. H., 1995: Interhemispheric correlation of late Pleistocene glacial events, Science, 269, 1541-1549.

Lyle, M., D. W. Murray, B. P. Finney, J. Dymond, J. M. Robbins, and K. Brooksforce, 1988: The record of late Pleistocene biogenic sedimentation in the eastern tropical Pacific Ocean, Paleoceanography, 3, 39-59.

Lyle, M., F. G. Prahl, and M. A. Sparrow, Upwelling And Productivity Changes Inferred from a Temperature Record in The Central Equatorial Pacific, Nature, 355, 812-815, 1992.

MacAyeal, D. R., 1993a: A Low-Order Model of the Heinrich Event Cycle. Paleoceanography, 8, 767-773.

MacAyeal, D. R., 1993b: Binge/Purge oscillations of the Lauerentide Ice Sheet as a cause of the North Atlantic's Heinrich Events, Paleoceanography, 8, 775-784.

Martinson, D. G., Pisias, N. G., Hays, J. D., Imbrie, J., Moore, T. C. Jr., and Shackleton, N. J., 1987: Age

dating and the orbital theory of ice ages: development of a high-resolution 0 to 300,000 years chronostratigraphy, Quaternary Research, 27, 1-29.

Maslin, M. A., Shackleton, N. J., and Pflaumann, U., 1995: Surface water temprature, salinity and density changes in the northeast Atrlantic during the last 45,000 years: Heinrich events, deep-water formation, and climatic rebounds. Paleoceanography, 10, 527-543.

Maslin, M., and Tzedakis, C., 1996: Sultry Last Interglacial Gets a Sudden Chill. Earth in Space, Vol. 9, No. 7, March 1997, p12-14

Maslin, M. A., Thomas, E., Shackleton, N. J., Hall, M. A., and Seidov, D., 1997: Glacial Northeast Atlantic surface water pCO2: productivity and deep-water formation. Marine Geology, 144: 177-190

Mayewski, P. A., Meeker, L. D., Twickler, M. S., Whitlow, S., Yang, Q., Lyons, W. B., and Prentice, M., 1997: Major features and forcing of high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series. Journal of Geophysical Research, v/. 102, 26,345-26,365

McCabe, A. M., and Clark, P. U., 1998: Ice-sheet variability around the North Atlantic Ocean druing the last deglaciation, Nature, 392, 373-377 McManus, J. F., Bond, G. C., Broecker, W. S., Johnsen, S., Labeyrie, L., and Higgins, S., 1994, High resolution climate records from the North Atlantic during the last interglacial, Nature, 371, 326-329

Meeker, L. D., Mayewski, P. A., Twickler, M. S., Whitlow, S. I., and Meese, D., 1997: A 110,000 year history of change in continental biogeniuc emissions and related atmospheric circulation inferred from the Greenland Ice Sheet Project Ice Core, Journal of Geophysical Research, 102, 26,489-26,505

Mikolajewicz, U., and Maier-Reimer, E., 1994: Mixed boundary conditions in OGCMs and their influence on the stability of the model's conveyor belt. Journal of Geophysical Research, 99, 22,633-22,644.

Mudelse, M., and Schulz, M., 1997: The Mid-Pleistocene climate transition: onset of 100 ka cycle lags ice volume build up by 280 ka. Earth and Planetary Science Letters, 151, 117-123

Muller, R. A., and MacDomald, G. J., 1997: Science, 277, 215-218

O'Brien, S. R., Mayewski, A., Meeker, L. D., Meese, D. A., Twickler, M. S., and Whitlow, S. I., 1996: Complexity of Holocene climate as reconstructed from a Greenland ice core. Science, 270, 1962-1964

Oppo, D. W., Horowitz, M., and Lehman, S. J., 1997: Marine evidence for reduced deep water production during Termination II followed by a relatively stable substage 5e (Eemian), Paleoceanography, 12 51-63

Oppo, D.W., J. F. McManus, J. L. Cullen, 1998: Abrupt Climate Events 500,000 to 340,000 Years Ago: Evidence from Subpolar North Atlantic Sediments. Science, 279 1335-1338.

Overpeck, J., Hughen, K., Hardy, D., Bradely, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A., and Zielinski, G., 1997: Arctic environmental change of the last four centuries, Science, 278, 1251-1256

Peglar S.M. 1993: The mid-Holocene Ulmus decline at Diss Mere, Norfolk, UK: a year-by-year pollen stratigraphy from annual laminations. The Holocene, 3 1-13.

Peteet, D. M., 1995: Global Younger Dryas, Quaternary International, 28, 93-104.

Petit et al. 1997: TITLE. Nature 387, 359-360

Pewe, T. L., Berger, G. W., Westgate, J. A., Brown, P. M., and Leavitt, S. W., 1997: Eva Interglaciation Forest Bed, Unglaciated East-Central Alaska: Global warming 125,000 Years Ago, Geological Society of America, Spec. Pa, 319, 317 p

Rackham O. 1980: Ancient Woodland: its history, vegetation and uses in England. Arnold, London

Rahmstorf, S., 1994: Rapid climate transitions in a coupled ocean-atmosphre model. Nature, 372, 82-85

Rahmstorf, S., Marotzke, J., and Willebrand, J., 1996: Stability of the Thermohaline Circulation, In: The Warmwatersphere of the North Atlantic Ocean, W. Kraus, ed., Gebrueder Borntraeger, Berlin, 129-157

Rasmussen T.L., van Weering T.C.E. & Labeyrie L. 1997: Climatic instability, ice sheets and ocean dynamics at high northern latitudes during the last glacial period (58-10 KA BP). Quaternary Science Reviews 16, 73-80.

Raymo et al. 1998: TITLE. Nature April 15th

Revel, M., Sinko, J. A., Grousset, F. E., and Biscaye, P., 1996: Sr and Nd isotopes as tracers of North Atlantic lithic particles: paleoclimatic implications. Paleoceanography, 11, 95-113.

Ruddiman, W. F., 1977: Late Quaternary deposition of ice-rafted sand in the subpolar North Atlantic (lat. 40o to 65oN). Geological Society of America, Bull., 88, 1813-1827.

Rind, D. and Chandler, M., 1991: Increased ocean heat transports nd warmer climate. Journal of Geophysical Research, 96, 7437-7461.

Rind, D., and Overpeck, J., 1993: Hypothesized causes of decade-to-century-scale climate variability: climate model results, Quaternary Science Reviews, 12, 357-374.

Sarnthein, M. and Tiedemann, R., 1990: Younger Dryas-style cooling events at glacial terminations I-VI at ODP Site 685: associated benthic d13C anomalies constrain meltwater hypothesis, Paleoceanography, 5, 1041-1055.

Schulz H., von Rad, U. and Erlenkeuser H., 1998: Correlation betwen Arabian Sea and Greenland Climate oscillations of the past 110,000 years. Nature, 393, 54-57.

Seidenkrantz, M.-S., 1993: Benthic firaminiferal and stable isotope evidence for a "Younger Dryas-style" cold spell at the Saalian-Eemian transition, Denmark, Palaeogeography, Palaeoclimatology, Palaeoecology, 102, 103-120

Seidenkrantz, M.-S., Kristensen, P., and Knudesn, K. L., 1995: Marine evidence for climatic instability during the last interglacial in shelf records from northwest Europe. J. Quat. Sci. 10 , 77-82 Seidenkrantz, M.-S., Bornmalm, L., Johnsen, S. J., Knudsen, K. L., Kuijpers, A., Lauritzen, S.-E., Leroy, S. A. ., Mergeai, I., Schweger, C., and van Vlet-Lanoe, B., 1996. Two-step deglaciation at the oxygen isotope stage 6/5e transition: the Zeifen-Kattegat climate oscillation. Quaternary Science Reviews, 15, 63-75

Seidov and Maslin, 1996: REFERENCE NEEDED

Severinghaus. J. P., Sowers, T., Brook, E. J., Alley, R. B., and Bender, M. L., 1998: Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice, Nature 391, 141-146

Shackleton 1969: The last interglacial in the marine and terrestrial records, Proc. Roy. Soc. London, B 174, 135-154

Shaffer and Bendtsen, 1994: REFERENCE NEEDED

Sirocko, F., Sarnthein, M., Erelnkeuser, H., Lange, H., Arnold, M., and Duplessy, J.-C., 1993: Century-scale events in monsoonal climate over the past 24,000 years. Nature, 364, 322-324.

Slowey, N. C., Henderson, G. M., and Curry, W. B., 1996: Direct U-Th dating of marine sediments from the two most recent interglacial periods. Nature, 383, 242-244.

Sowers, T., Bender, M., Labeyrie, L., Martinson, D., Jouzel, J., Raynaud, D., Pichon, J. J., and Korotevich, Y. S., 1993: A 135,000 year Vostok-SPECMAP common temporal framework. Paleoceanography 8, 737-764

Stirling, C. H., Esat, T. M., McCulloch, M. T., and Lambeck, K., 1995: High-precision U-series dating of corals from Western Australia and implications for the timing and duration of the last interglacial. Earth and Planetary Science Letters, 135, 115-130.

Szabo, B. J., Ludwig, K. R., Muhs, D. R., and Simmons, K. R., 1994: Thorium-230 ages of corals and duration of last interglacial sea-level hgihsatnd on Oaho, Hawaii, Science, 266, 93-96.

Taylor, K.C., Alley, R. B., Doyle, G. A., Grootes, P. M., Mayewski, P. A., Lamorey, G. W., White, J. W. C., and Barlow, L. K., 1993: The 'flickering switch' of late Pleistocene climate change. Nature 361, 432-436.

Taylor, K. C., P. A. Mayewski, R. B. Alley, E. J. Brook, A. J. Gow, P. M. Grootes, D. A. Meese, E. S. Saltzman, J. P. Severinghaus, M. S. Twickler, J. W. C. White, S. Whitlow, G. A. Zielinski 1997: The Holocene-Younger Dryas Transition Recorded at Summit, Greenland. Science, 278 825-827.

Thomas, E., Booth, L., Maslin, M., and Shackleton, N. J., 1995: Northeastern Atlantic benthic foraminifera during the last 45,000 years: productivity changes as seen from the bottom u Paleoceanography, 10, 545-562

Thompson, L. G., 1989: Ice-Core Records with Emphasis on the global record of the last 2000 years. In: Global Changes of Past, R. Bradley, ed., UCAR/Office of Interdisciplinary Earth Studies, Boulder, CO (USA), Vol. 2, 201-223.

Thompson, L. G., and E. Mosley-Thompson. 1987: Evidence of abrupt climatic damage during the last 1,500 years recorded in ice cores from the tropical Quelccaya ice cap, Peru. P 99-110 in Abrupt Climatic Change--Evidence and Implications, W. H. Berger and L. D. Labeyrie, eds.

Thompson, L. G., Mosley-Thompson, E., Davis, M. E., Bolzan, J. F,. Dai, J., Yao, T., Gundestrup, N., Wu, X., Lein, L., and Xie, Z., 1989: Holocene - late Pleistocene climatic ice core records from Qinghai-Tibetan Plateau. Science, 246, 474-477.

Thompson, L. G., Mosley-Thompson, E., Davis, M., Lin, P. N., Yao, T., Dyurgerov, M., and Dai, J., 1993: "Recent warming": ice core evidence from tropical ice cores with emphasis on Central Asia. Global and Planetary Change, 7, 145-156.

Thompson, L. G., Mosley-Thompson, E., Davis, M. E., Lin, P.-N., Henderson, K. A., Cole-Dai, J., Bolzan, J. F., and Liu, K.-B., 1995: Late glacial stage and Holocene tropical ice core records from Huascarn, Peru: Science, 269, 46-50.

Thompson, L. G., E. Mosley-Thompson, W. Dansgasrd, and P. M. Grootes. 1986: The "Little Ice Age" as recorded in the stratigraphy of the tropical Quelccaya ice ca Science 234:361-364.

Thouveny, N., de Beaulieu, J. L., Bonifay, E., Creer, K. M., Guiot, J., Icole, M., Johnsen, S., Jouzel, J., Reille, M., Williams, T., and Williamson, D., 1994: Climate variations in Europe over the past 140 kyr deduced from rock magnetism, Nature 371, 503-506

Tzedakis, P. C., Andrieu, V., de Beaulieu, J.-L., Crowhurst, S., Folieri, M., Hooghiemstra, H., Magri, D., Reille, M., Sadori, L., Shackleton, N. J., and Wijmstra, T. A., 1997: Comparison of terrestrial and marine records of changing climate of the last 500,000 years. Earth and Planetary Science Letters, 150, 171-176

Van Andel T.H. & Tzedakis P.C. 1996: Palaeolithic landscapes of Europe and environs: 150,000-25,000 years ago: an overview. Quaternary Science Reviews. 15 481-500.

Winograd, I. J., Szabo, B. J., Coplen, T. B., and Riggs, A. C., 1988: A 250,000-year climatic record from Great Basin vein calcite: implications for Milankovich theory. Science, 242, 1275-1280.

Winograd, I. J., Coplen, T. B., Landwehr, J. M., Riggs, A. C., Ludwig, K. R., Szabo, B. J., Kolesar, P. T., and Revesz, K. M., 1992: Continuous 500,000-year climate record from vein calcite in Devil's Hole, Nevada, Science, 258, 255-260.

Winograd I. J., Landwehr J. M., Ludwig K. R., Coplan T. B. and Riggs A. C., 1997: Duration and structure of the past four interglaciations. Quaternary Research, 48, 141-154.

Woillard, G., 1978: Grande Pile peat bog: a continuous pollen record for the last 140,000 years. Quaternary Research, 12, 152-155.

Woillard and Mook, 1982: REFERENCE Nature

Wright H.E. jr 1993: Environmental determinism in Near Eastern Prehistory. Current Anthropology 34, 458-469.

Yiou, P., Fuhrer, K., Meeker, L. ., Jouzel, J., Johnsen, S., and Mayewski, P. A., 1997: Paleoclimatic variability inferred from the spectral analysis of Greenland and Antarctic ice core data. Journal of Geophysical Research, 102, 26,441-26, 453.

Zagwijn, W. H., 1963: Pleistocene stratigraphy in the Netherlands, based on changes in vegetation and climate, Ned. Geol. Mijnbouwkundig Genootschap, Verhandelingen, Geologische Serie, 21-2, 173-196

Zagwijn, W. H., 1975: Variations in climate as shown by pollen analysis, especially in the lower Pleistocene of Europe. In: Ice Ages: Ancient and Modern, A. E. Wright and F. Moseley, eds., Seel House Press, Liverpool, UK, 137-152

Zhisheng A. & Porter S.C. 1997: Millenial-scale climatic oscillations during the last interglaciation in central China. Geology 25 603-606.

Tables for this paper

Table 1 (Ages in calender years after Bond et al., 1997 for H0- H3; after Bond et al., 1993 for H4- H6).

Timing of major Heinrich events during the last 130,000 years

YD or H0 12.2 ka (calendar age)

H1 16.8

H2 24.1

H3 30.1

H4 35.9

H5 50.0

H6 66.0

YD: Younger Dryas

H: Heinrich event

Table 2 . The time scale of the last 130,000 years since the start of the Eemian warm period. Ages of isotope stages after Martinson et al. (1987) in the first column, after Imbrie et al 1984 in the second column. [ ] Event Age (ka)

last deglaciation (event 2.0), termination i: 12,050 yr [12]

glacial maximim (event 2.2): 17,850 yr [19]

boundary stage 2/3 (event 3.0): 24,110 yrs [24]

boundary stage 3/4 (event 4.0): 58,960 yrs [59]

boundary stage 4/5 (event 5.0): 73,910 years [71]

event 5.1: 79,250 kyr [80]

event 5.2: 90.950 kyr [87]

event 5.3: 99,380 kyr [99]

event 5.4: 110.790 kyr [107]

event 5.5: 123,820 kyr [122].

boundary stage 5/6 9event 6.0, termination ii): 129,840 yrs [128]

Note: Stage 5 is defined as an interglacial stage that contains 3 negative events (i.e., warm), labeled 5.1, 5.3, and 5.5; and two positive events - cold - labeled 5.2 and 5.4. this numbering tends to supersede the 5a through 5 e names.

List of Figures for this paper.
